Lecture 25. Radiation in energy balance models

<u>Objectives:</u>

1. A hierarchy of climate models.

2. Examples of simple energy balance models.

<u>Required reading</u>:

L02: 8.5

Radiative Equilibrium and Energy Balance models

Radiative Equilibrium: occurs when the total (shortwave + longwave) radiative heating is zero at all levels.

Energy balance models solve for planetary temperatures by balancing radiative energy (and perhaps other energy transports).

There is a hierarchy of energy balance models of differing complexity.
These models:

1) can explain temperature profiles in planetary atmospheres.
2) are used to analyze climate sensitivity to radiative forcing.

1. A hierarchy of climate models

- Climate models may be classified by their dimensions:

Zero Dimensional Models (0-D):

consider the Earth as a whole (no change by latitude, longitude, or height)

One Dimensional Models (1-D):

allow for variation in one direction only (e.g., resolve the Earth into latitudinal zones or by height above the surface of the Earth)

Two Dimensional Models (2-D):

allow for variation in two directions at once (e.g., by latitude and by height)

Three Dimensional Models (3-D)

allow for variation in three directions at once (i.e., divide the earth-atmosphere system into domains, each domain having its own independent set of values for each of the climate parameters used in the model.

- Climate models may be classified by the basic physical processes included into the consideration:

Energy Balance Models:

0-D or 1-D models (e.g., allow to change the albedo by latitude) calculate a balance between the incoming and outgoing radiation of the planet;

Radiative Convective Models (see Lecture 26):

1-D models to model the temperature profile the atmosphere by considering radiative and convective energy transport up through the atmosphere.

General Circulation Climate Models:

2-D (longitude-averaged) or 3-D climate models solve a series of equations and have the potential to model the atmosphere very closely.

2. Examples of simple energy balance models.

Recall lecture 24:

$$\frac{F_0}{4}(1-\overline{r}) = F_{IR}^{\uparrow TOA}$$

"Bare Earth" model: no atmosphere, surface absorbs all solar radiation and emits IR as a blackbody

$$\Downarrow$$

T_e = 255 K = -18^{0}C is very low!!!

NOTE: $\mathbf{T_e}$ is much less than the global average surface temperature, $\mathbf{T_s}$ (about 288 K) due to greenhouse effect.

How would we measure the greenhouse effect?

Upwelling flux at the surface: $F_s^{\uparrow} = \sigma T_s^4$

Upwelling longwave flux $F^{\uparrow}$ at TOA – from satellite

The difference between the upwelling fluxes at the surface and TOA gives the greenhouse effect:

$$G = \sigma T_s^4 - F^{\uparrow} \qquad [25.1]$$

This is the amount of heat (measured in Watts) per unit area of the Earth.

What is a reasonable estimate of the greenhouse effect?

$$F^{\uparrow} = 235 \text{ W m}^{-2}$$
$$T_s = 288K$$
$$\sigma T_s^4 = 390 \text{ W m}^{-2}$$
$$G = 390 - 235 = 155 \text{ W m}^{-2}$$

Simple model of greenhouse effect #1: single layer gray energy balance model

Let's include the atmosphere assuming that it emits (absorbs) as a gray body. Assume that the atmosphere does not absorb solar radiation - all is absorbed at the surface.

Assumptions: 1) Isothermal layer, 2) Gray atmosphere in LW,
3) Black surface in LW, 4) No atmosphere absorption in SW.

Absorbed solar flux is $$F_{sun}^{\downarrow} = \frac{F_0}{4}(1-\bar{r}) = \sigma T_e^4$$

F_0 is solar constant, r is mean albedo, T_e is effective temperature,
ε_a is LW atmosphere emissivity

TOA balance: $F_{sun} = F_{atm}^{\uparrow} + (1-\epsilon_a)F_{sfc}^{\uparrow}$

$$\frac{F_0}{4}(1-\bar{r}) = \varepsilon_a \sigma T_a^4 + (1-\varepsilon_a)\sigma T_s^4 \qquad [25.2]$$

Surface balance: $F_{sun}^{\downarrow} + F_{atm}^{\downarrow} = F_{sfc}^{\uparrow}$

$$\frac{F_0}{4}(1-\bar{r}) + \varepsilon_a \sigma T_a^4 = \varepsilon_a \sigma T_s^4 \qquad [25.3]$$

Two unknowns: surface and atmosphere temperature - T_s and T_a

Eliminate $F^{\uparrow}_{sfc}$: $F_{atm} = \frac{\epsilon_a}{2-\epsilon_a} F_{sun}$

Eliminate $F^{\downarrow}_{atm}$: $F^{\uparrow}_{sfc} = \dfrac{2F^{\downarrow}_{sun}}{(2-\varepsilon_a)}$

we obtain surface temperature:

$$T_s^4 = \frac{F_0(1-\bar{r})}{2\sigma(2-\varepsilon_a)} \qquad [25.4]$$

for ε = 0.6 => $\mathbf{T_s}$ =278 K

NOTE: increasing for ε increases $\mathbf{T_s}$. This is so-called "**runaway greenhouse effect**" :

warmer Ts => more evaporation => more water vapor => higher emissivity =>warmer Ts

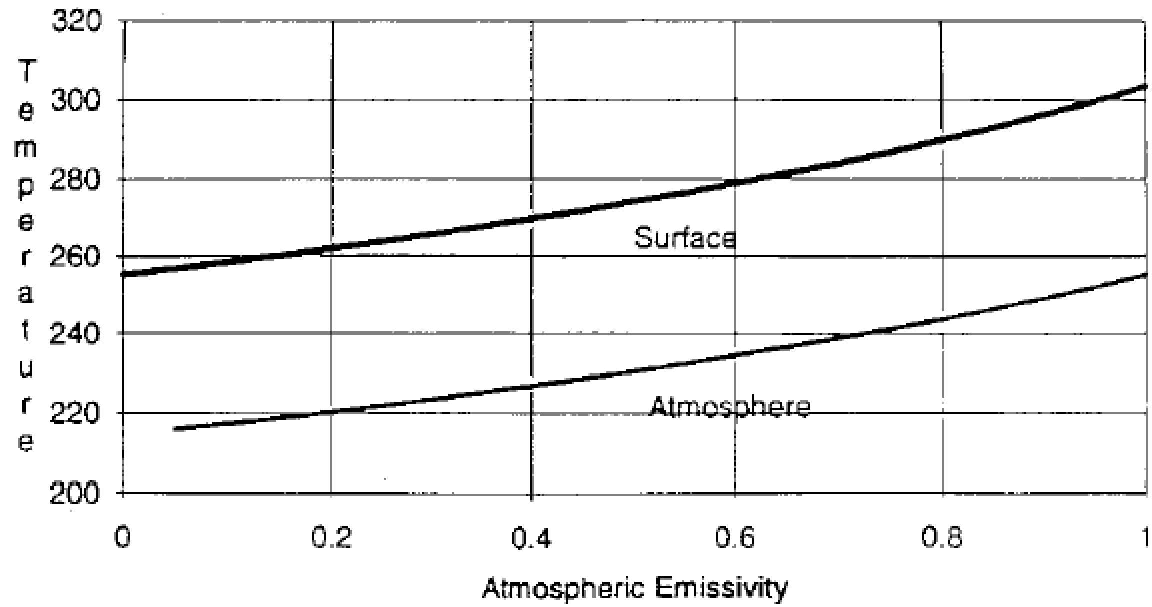
Why? Because we assume that atmosphere is a gray body.

In terms of temperatures:

$$T_a = \frac{T_e}{(2-\epsilon_a)^{1/4}} \leq T_e \qquad T_s = \frac{T_e}{(1-\epsilon_a/2)^{1/4}} = 2^{1/4} T_a \geq T_e$$

Limits: thin atmosphere $\epsilon_a \to 0 \;\Rightarrow\; T_a = T_e/2^{1/4} \quad T_s = T_e$

black atmosphere $\epsilon_a = 1 \;\Rightarrow\; T_a = T_e \quad T_s = 2^{1/4} T_e$



One-layer gray model temperatures as a function of atmospheric emissivitivy for the Earth's globally averaged absorbed solar energy (239 W/m^2).

Simple model of greenhouse effect #2: single black layer with spectral window energy balance model:

A more realistic way to deal with partial longwave transparency of the atmosphere is to assume that a fraction of the spectrum is clear.

Assume that

$f = \frac{1}{\sigma T^4} \int_{\nu_1}^{\nu_2} B_\nu(T) d\nu$ is the fraction of LW spectrum which is completely transparent

and the remainder of the LW spectrum is black. Surface is black and no SW absorption by the atmosphere.

TOA balance: $$\frac{F_0}{4}(1-\bar{r}) = f\sigma T_s^4 + (1-f)\sigma T_a^4 \qquad [25.5]$$

Surface balance: $$\frac{F_0}{4}(1-\bar{r}) + (1-f)\sigma T_a^4 = \sigma T_s^4 \qquad [25.6]$$

Thus we can express $\mathbf{T_s}$ and $\mathbf{T_a}$ via $\mathbf{T_e}$

$$T_s = \left(\frac{2}{1+f}\right)^{1/4} T_e \qquad \text{and} \qquad T_a = \left(\frac{1}{1+f}\right)^{1/4} T_e \qquad [25.7]$$

Limits:

No window f=0 => $T_s = 2^{1/4} T_e$ and $T_a = T_e$

All window f=1 => $T_s = T_e$ and $T_a = T_e / 2^{1/4}$

Earth: f is about 0.3 => T_s = 284 K and T_a = 239 K

<u>How to make the model more realistic:</u>

Tropics: radiation excess

North poles and high latitudes: radiation deficit

$$\Downarrow$$

must be poleward transport of energy and upward (surface to atmosphere) transport of energy

Multiple black layers with spectral window

Need multiple atmosphere layers to explain atmosphere temperature profile.

Black layers $\rightarrow$ LW flux absorbed by a layer comes from adjacent layers.
Portion of surface flux emitted in window escapes directly to space.

TOA: $F_{sun} = \sigma T_e^4 \; = \; (1-f)\sigma T_1^4 + f\sigma T_s^4$
Layer 1: $(1-f)\sigma T_2^4 = 2(1-f)\sigma T_1^4$
Layer 2: $(1-f)\sigma T_1^4 + (1-f)\sigma T_3^4 = 2(1-f)\sigma T_2^4$
Layer n: $(1-f)\sigma T_{n-1}^4 + (1-f)\sigma T_{n+1}^4 = 2(1-f)\sigma T_n^4$
Surface: $F_{sun} + (1-f)\sigma T_N^4 = \sigma T_s^4$

Cancel $1-f$ factors: $T_{n+1}^4 = 2T_n^4 - T_{n-1}^4 \;\; \rightarrow \;\; T_N^4 = NT_1^4$
Two unknowns T_1 and T_s; solve to get

$$T_s = T_e \left(\frac{1+N}{1+fN}\right)^{1/4} \qquad T_1 = T_e \left(\frac{1}{1+fN}\right)^{1/4}$$

Thicker atmosphere (larger N) gives warmer surface and cooler TOA
$\rightarrow$ greater greenhouse effect.
Spectral window ($f > 0$) allows top of atmosphere to be colder than T_e.

One-dimensional (latitude) energy balance model:

(Budyko 1969; Sellers 1969, Cess 1976)

Atmosphere is only implicit: TOA outgoing longwave flux is parameterized as a function of T_s

Budyko's parameterization is based on monthly mean atmospheric temperature and humidity profiles, and cloud cover observed at 260 stations

$$F_{LW}(x) = a_1 + b_1 T_s(x) - [a_2 + b_2 T(x)]\eta \qquad [25.8]$$

where a_i and b_i are the empirical constants based on statistical fitting, x =sin (ϕ) and ϕ is latitude. If cloud cover is taken constant of 0.5 then

$$F_{LW}(x) = (1.55W/m^2/K)T_s(x) - 212W/m^2 \qquad [25.9]$$

NOTE: The approximation for linear relation between OLR and the surface temperature may be argued from the fact that the temperature profiles have more or less the same shape at all latitudes, and that OLR, which depend on temperatures at all levels, may be expressed as a function of surface temperature.

Annual mean TOA solar insolation fit well with

$$S(x) = F_0 / 4[1 - 0.482 P_2(x)] \qquad [25.10]$$

$P_2(x) = (3x^2-1)/2$ is the second Legendre polynomial.

Thus energy balance equilibrium ($\delta T/\delta t=0$) with diffuse transport:

$$-D\frac{\partial}{\partial x}(1-x^2)\frac{\partial F_{LW}}{\partial x} + F_{LW} = S(x)[1-r(x)] \qquad [25.11]$$

where D is diffusion coefficient for energy transport and r(x) is albedo.

With no meridional transport, the poles are way too cold.

These models have been used to study the ice-albedo feedback by having the albedo $r(x)$ depend on temperature $T(x)$.

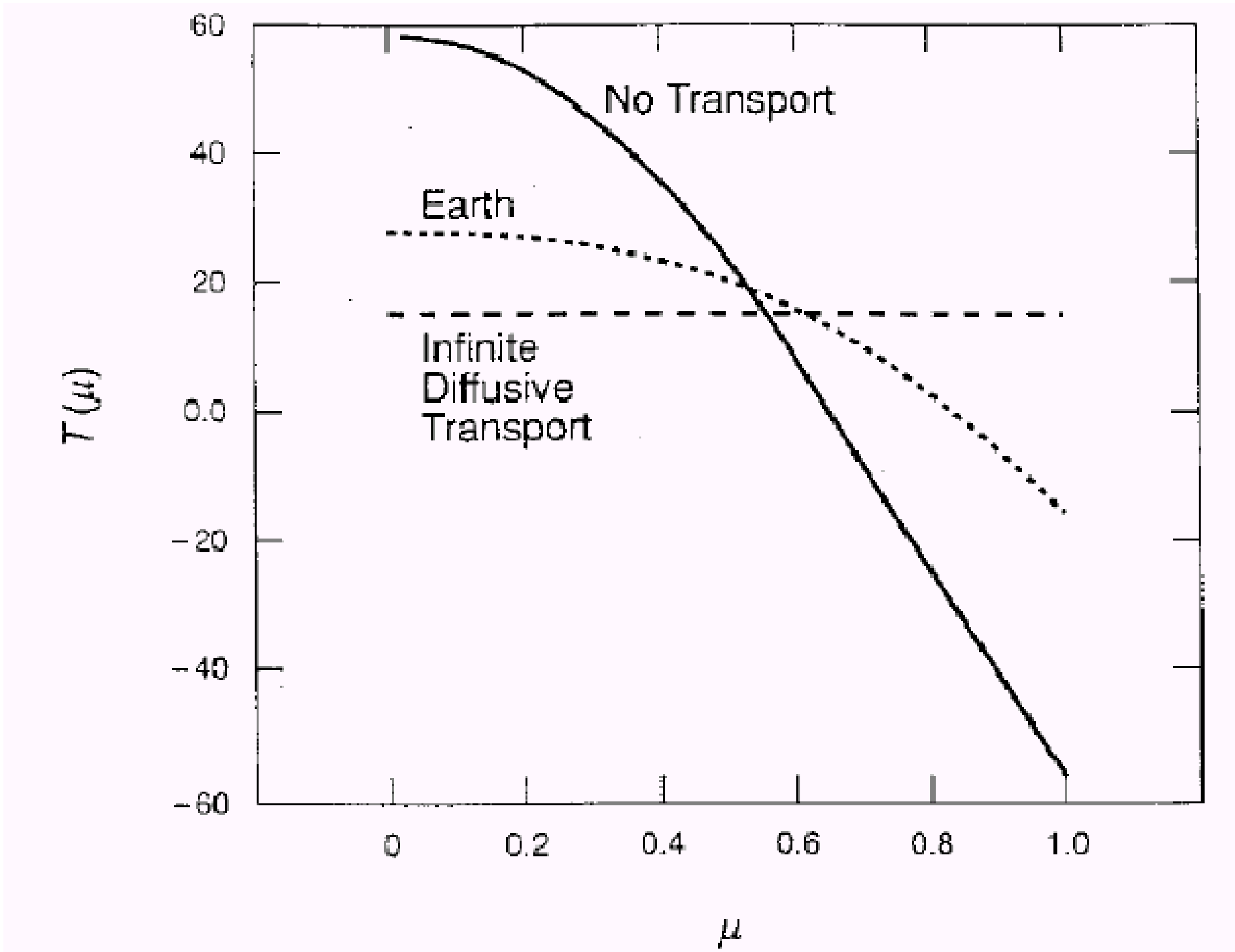
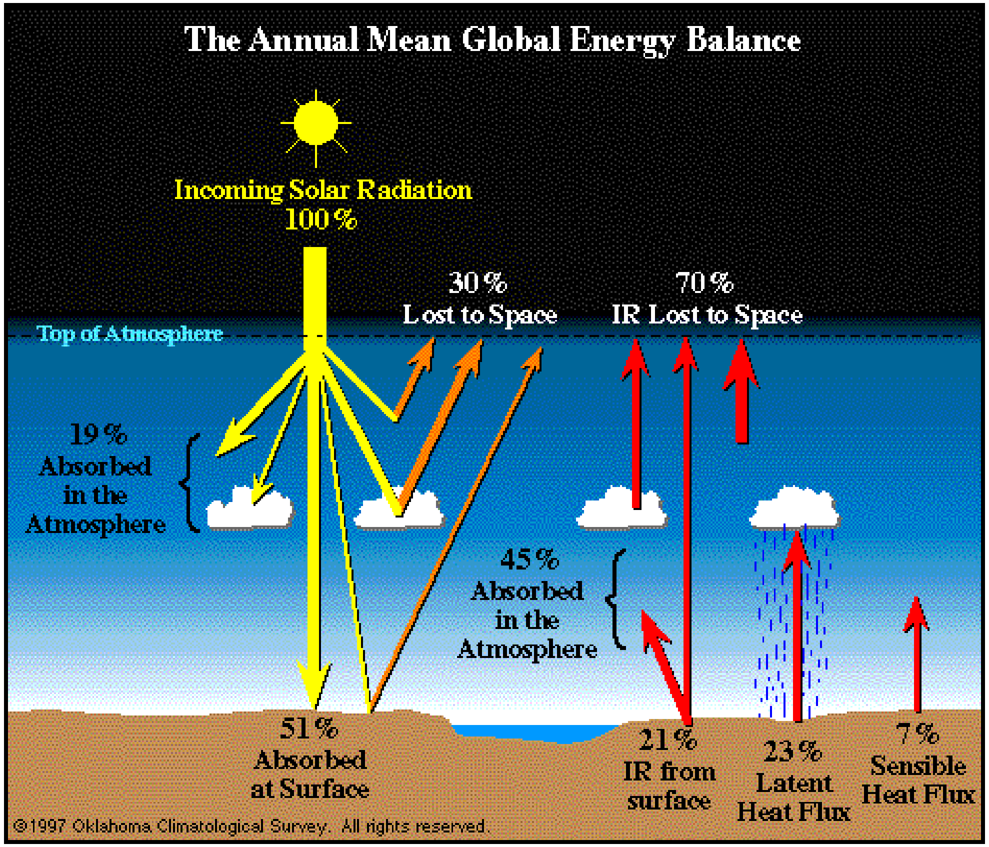


Figure 25.1 Zonally average surface temperature (K) as a function of the sine of the latitude, μ (μ is same as x in [25.11]), observed and for cases of no horizontal heat transport, and infinite horizontal heat transport (from North et al., 1981).

How to make the model more realistic:

In addition to radiative energy, account for latent and sensible heat



Absorption and re-emission of radiation at the earth's surface is only one part of an intricate web of heat transfer in the earth's planetary domain. Equally important are selective absorption and emission of radiation from molecules in the atmosphere. If the earth did not have an atmosphere, surface temperatures would be too cold to sustain life. If too many gases which absorb and emit infrared radiation were present in the atmosphere, surface temperatures would be too hot to sustain life. (http://okfirst.ocs.ou.edu/train/meteorology/EnergyBudget2.html)

Figure 25.2 Simplify representation of energy flow in the earth-atmosphere climate system.

Global energy balance

- When averaged over a year, the incoming energy in both the earth and its atmosphere equals the outgoing energy. If we consider the entire Earth-atmosphere system, then the amount of radiation entering the system must equal to the amount leaving, or the system would continually heat or cool. Not all of this energy is radiative energy; some is sensible and latent heat.
- **Latent Heat -** the heat released or absorbed by a substance during a phase change
- **Sensible Heat -** the heat absorbed or transmitted when the temperature of a substance changes but the substance does not change state
- If we consider the atmosphere alone, we find that the atmosphere experiences radiative cooling. The atmosphere is kept from a net cooling by the addition of energy by latent and sensible heating.
- The atmosphere has a warming effect on Earth's surface -- the "atmospheric greenhouse effect". If Earth had no atmosphere, the globally averaged surface temperature would be -18 degrees Celsius. Because Earth does have an atmosphere, the average surface temperature actually is 15 degrees Celsius.

Continued

- The atmosphere acts as a greenhouse because of gases that selectively allow solar radiation to pass through but absorb and then re-emit terrestrial radiation. These gases are collectively called "greenhouse gases" and include water vapor, carbon dioxide, ozone, molecular oxygen, methane and nitrous oxide. These gases are selective as to which wavelengths they will absorb. For example, ozone absorbs shortwave ultraviolet radiation whereas water vapor absorbs infrared radiation more readily.
- Most of the sun's radiation that passes through the atmosphere to hit the earth is in the visible part of the spectrum.
- Most of the earth's radiation that escapes the atmosphere is in the infrared band between 8 microns and 11 microns. This region of the spectrum is called the "atmospheric window".

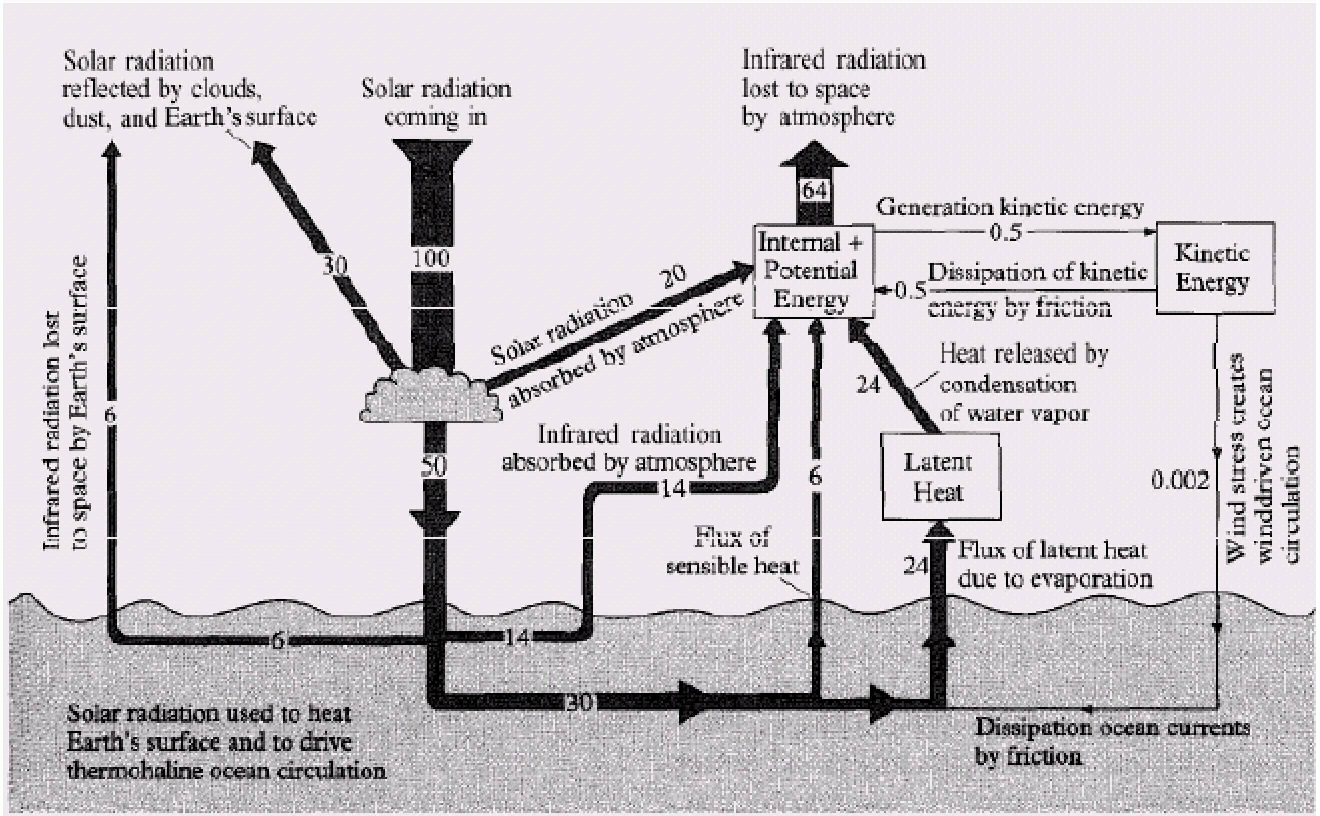


Figure 25.3 Schematic diagram of energy flow in the earth-atmosphere climate system (from Piexoto and Oort, 1992).